

Terrestrial Conditions at the Last Glacial Maximum and CLIMAP Sea-Surface Temperature Estimates: Are They Consistent?

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Received April 13, 1984

CLIMAP (1981, "Seasonal Reconstruction of the Earth's Surface at the Last Glacial Maximum," Geological Society of America Map and Chart Series MC-36) boundary conditions were used as inputs to the GISS general circulation model, and the last glacial maximum (LGM) climate was simulated for six model years. The simulation was compared with snow line depression and pollen-inferred temperature data at low latitudes, specifically for Hawaii, Colombia, East Africa, and New Guinea. The model does not produce as much cooling at low latitudes as is implied by the terrestrial evidence. An alternative experiment in which the CLIMAP sea-surface temperatures were uniformly lowered by 2°C produces a better fit to the land data although in Hawaii model temperatures are still too warm. The relatively warm CLIMAP tropical sea-surface temperatures also provide for only a slight decrease in the hydrologic cycle in the model, in contrast to both evidence of LGM tropical aridity and the results of the experiment with colder ocean temperatures. With the CLIMAP sea-surface temperatures, the LGM global annual mean surface air temperature is 3.6°C colder than at present; if the ocean temperatures were allowed to cool in conformity with the model's radiation balance, the LGM simulation would be 5°-6°C colder than today, and in better agreement with the tropical land evidence. © 1985 University of Washington.

INTRODUCTION

Until about 1960 it was thought that the equatorial regions during glacial periods were either essentially unaffected or experienced a pluvial climate (Flenley, 1979). Recent geological investigation of late Pleistocene tropical glacial advances, however, is consistent with emerging palynological and lake-level data suggesting temperature depression and aridity at low latitudes during the last glacial maximum (LGM).

The CLIMAP (1981) sea-surface temperature reconstruction for the LGM approximately 18,000 yr B.P. reveals that the "ice-age ocean was strikingly similar to the present ocean in at least one respect: large areas of the tropics and subtropics within all oceans had sea-surface temperatures as warm as, or slightly warmer than, today" (CLIMAP Project Members, 1981, p. 9).

Are these two lines of evidence consis-

tent? Is it possible that warm ocean surface waters and substantial continental temperature depression were both characteristic of the LGM climate in low latitudes? Webster and Stretten (1978) raised this question in connection with the snow line depression in New Guinea, but it applies essentially throughout the tropics [e.g., the list of tropical and semitropical LGM mountain glaciers of Hastenrath (1981, Table 5) and Denton and Hughes (1981)]. To explore this issue we have incorporated the CLIMAP sea-surface temperatures into a general circulation model, along with other ice-age boundary conditions, and computed the resulting change in climate. Particular emphasis is on tropical areas with snow line and pollen evidence, and four regions are discussed in detail: Hawaii, Colombia, East Africa, and New Guinea. These cover a wide range of longitudes, including island locations and mountain chains on west and east coasts, at low and subtropical latitudes.

GLACIAL AND PALYNOLOGICAL EVIDENCE

The subtropical and tropical cooling indicated by the snow line and pollen data suggests a climatic change in widely dispersed geographic regions, including the four study areas (Table 1).

Snow line depression during the LGM on Mauna Kea, Hawaii (altitude 4206 m), is estimated at 935 ± 190 m (Porter, 1979). This last glaciation culminated sometime between $22,150 \pm 250$ and 9080 ± 200 yr B.P. (Porter, 1977a). Pollen studies, undertaken before the advent of radiocarbon dating, reveal a dominance of xerophytic subalpine elements prior to the postglacial interval which possibly are correlative with the last glaciation (Selling, 1948; Porter, 1979). Rain forest vegetation evidently was more restricted than at present, and tree line apparently was depressed to an altitude of about 2000 m (Porter, 1979), which is consistent with the inferred snow line depression.

Herd (1974) and Herd and Naeser (1974) recorded parallel evidence for a 950-m snow line depression before 13,800 yr B.P. in the Cordillera Central, Colombia. In summarizing the equatorial New World evidence for glaciation, Hastenrath (1981) described an early ice advance that extended down to a minimum of 3500 m possibly at some time after 25,000 yr B.P. Unfortunately, chronological control is lacking. Nearby in the Colombian sierra, Gonzales *et al.* (1965) described the last glaciation as culminating approximately 13,000 yr B.P. Van der Hammen (1974) reviewed pollen evidence demonstrating vegetational change in the late Pleistocene. Laguna de Fuquene (altitude 2580 m) provides a 12-m record spanning dates of $20,575 \pm 190$ and $10,820 \pm 60$ yr B.P. (Fig. 1a). At the LGM, open vegetation (Gramineae) dominated in place of forest, and at about 21,000 yr B.P. the inferred vegetation depression was approximately 1500 m. The extent of tree line lowering is summarized in pollen diagrams from the Colombian Andes (Fig. 1b). More

recent evidence from Laguna Ciega (altitude 3500 m) in the Colombian sierra also shows a very cold and dry period between bracketing dates of $20,840 \pm 140$ and $12,830 \pm 80$ yr B.P., with estimates of vegetation depression of 1200–1500 m (Van der Hammen *et al.*, 1981).

Using a method similar to that in Porter's (1979) study of snow line depression in Hawaii, Osmaston (1965) obtained a value of 900 m for snow line depression on Mt. Kilimanjaro in East Africa during the LGM. This glaciation has been dated between 30,000 and 10,000 yr B.P. (Downie and Wilkinson, 1972), and Hamilton (1982) has suggested that the glaciation was contemporaneous throughout the East African mountains. In his summary of glacial events in the tropics, Hastenrath (1981) also suggested probable glacial correlation between Mts. Kenya and Kilimanjaro, and the Ruwenzori. Several authors (Flenley, 1979; Livingstone, 1980; Hamilton, 1982) summarized the late Quaternary vegetational changes in the East African mountains: the more alpine elements were replaced by forest as deglaciation ensued (Fig. 2a). The LGM is bracketed by dates of greater than $12,890 \pm 130$ yr B.P. at Muchoya, between $33,350 \pm 1000$ and $14,050 \pm 360$ yr B.P. at Sacred Lake, between $27,750 \pm 600$ and $17,000 \pm 300$ yr B.P. at Kaisungor (Cherangani), and substantially below an interval at Rutundu dated $10,800 \pm 100$ yr B.P. Laboot Swamp, Mt. Elgon, Kenya (elevation 2880 m), recently investigated by Hamilton (1982), also shows distinctive vegetational changes between $23,073 \pm 120$ and $13,776 \pm 80$ yr B.P. Throughout East Africa, forest was greatly reduced in extent and considerable depression of the upper limit of vegetation zones is indicated.

The high mountains of Papua, New Guinea, were glaciated down to between 2750 and 3300 m (Löffler, 1972; 1975). This represents a minimum snow line depression of 900 m and is dated to before 12,600 yr B.P. on Mt. Wilhelm. Löffler (1977) refers to the accumulation of evidence suggesting

TABLE 1. SNOW LINE AND POLLEN DATA FOR THE LAST GLACIAL MAXIMUM

	Snow line depression (m)	Glacial-geological limiting dates for LGM (yr B.P.)	Vegetation-zone depression (m)	Pollen-indicated cooling interval (yr B.P.)	Pollen temperature depression estimates (°C)
Hawaii					
Mauna Koa	935 ± 190 Porter (1979)	22,000–9000 Porter (1977a)	900 Porter (1979)	?	“Cooler” Selling (1948)
Colombia					
Andes	950 Herd (1974) and Herd and Naeser (1974)	25,000–13,800 Hastenrath (1981), Herd (1974), and Herd and Naeser (1974)	1200–1500 Van der Hammen (1974) and Van der Hammen <i>et al.</i> (1981)	>21,000–10,820 Van der Hammen (1974) 20,840–12,830 Van der Hammen <i>et al.</i> (1981)	8–10 Van der Hammen (1974) and Flenley (1979)
East Africa					
Mt. Kilimanjaro, Mt. Kenya	900 Osmaston (1965)	30,000–10,000 Downie and Wilkinson (1972)	>700 Hamilton (1982)	>23,000–13,000 Hamilton (1982)	≥4.5 Hamilton (1982)
New Guinea					
Mt. Wilhelm	900 Löffler (1972, 1975)	?–15,000 Hope <i>et al.</i> (1976)	1200 Hope (1976) 1000–1500 Walker and Flenley (1979)	26,000–14,000 Hope (1976) and Walker and Flenley (1979)	2–8 (28,000–18,000 yr B.P.) 7–11 (18,000–16,000 yr B.P.) Flenley (1979) and Walker and Flenley (1979)

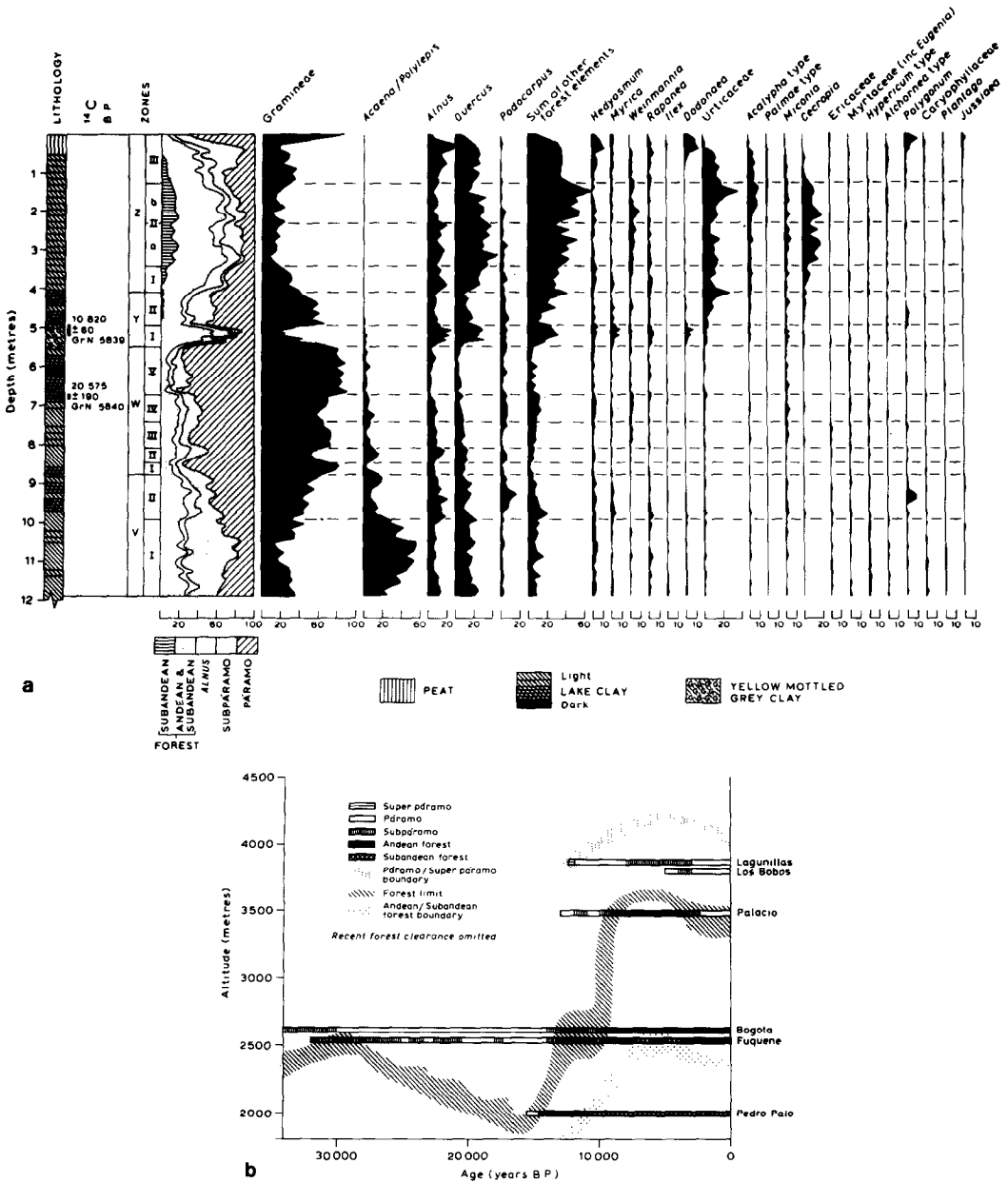


FIG. 1. (a) Pollen diagram from Laguna de Fuquene, Colombia (altitude 2580 m), covering the late Pleistocene (after Van der Hammen, 1974). (b) Summary diagram of late Quaternary vegetational changes in the Colombian Andes (Flenley, 1979).

contemporaneity between New Guinea glaciation and the Northern Hemisphere Wisconsin glacial advance that culminated approximately 15,000 yr B.P. (Hope *et al.*, 1976). The palynological evidence (Fig. 2b) suggests a depression of present vegetational zones by 1000–1500 m (Walker and

Flenley, 1979). Radiocarbon dates encompassing the LGM at Komanimambuno Mire are $21,760 \pm 350$ and $14,710 \pm 200$ yr B.P. (Hope, 1976); at Sirunki Swamp they are $25,800 \pm 650$, $19,900 \pm 650$, and $14,200 \pm 250$ yr B.P. (Walker and Flenley, 1979).

A summary of the snow line and vege-

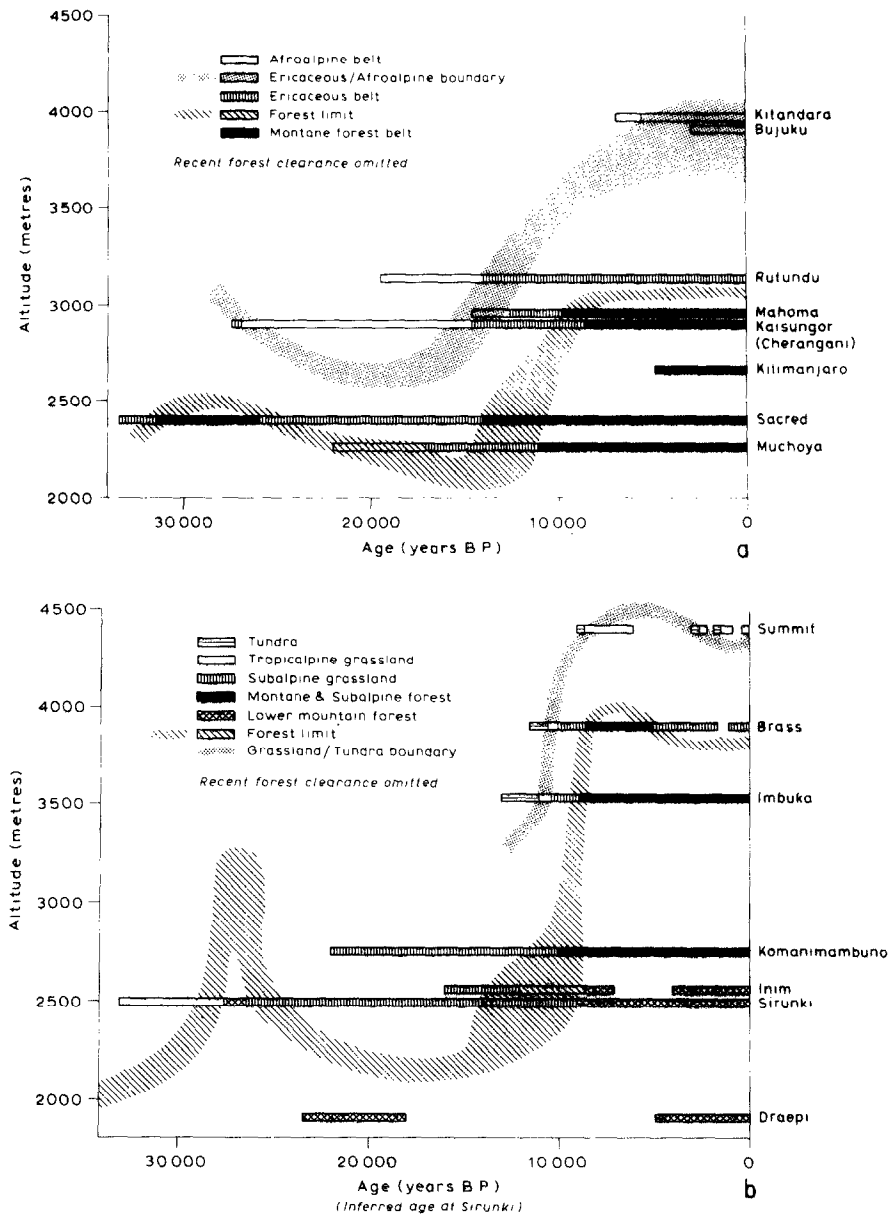


FIG. 2. (a) Summary diagram of late Quaternary vegetational changes in the East African mountains (Flenley, 1979). (b) Summary diagram of vegetational changes in the New Guinea Highlands (Hope, 1976; Flenley, 1979).

tational-zone depression is given in Table 1. There is a fairly consistent pattern indicating descent of about 1 km at these widely separated locations; as noted by Klute (1921), latitudinally averaged Pleistocene snow line depressions are also remarkably uniform in both the Northern and Southern Hemispheres. How does this

translate into indications of temperature change? Porter (1979) and Webster and Streten (1978) suggested that the snow line depression could result from increased precipitation in addition to temperature depression. The vegetational evidence for Colombia (Van der Hammen, 1974; Van der Hammen *et al.*, 1981), East Africa (Living-

stone, 1980; Palmer *et al.*, 1979), and New Guinea (Flenley, 1979) clearly indicates a cooler and drier climate. The explanation that satisfies both data sets is that the observed depression represents a cooling, with the freezing level and vegetation zones descending 1 km. Did the cooling occur at 18,000 yr B.P.? Both glacial and palynological ^{14}C dates span a wide interval; however, they are consistent (Table 1). While glacial evidence does not indicate the extent of the glaciers at 18,000 yr B.P., the vegetation zones are at or near their greatest depression at this time (Figs. 1 and 2).

Finally, assuming that the data indicate a temperature reduction, how can we calculate its magnitude? Estimates of temperature depression for the pollen data are given in Table 1 with the reference. They are derived by assuming a temperature change with altitude associated with the vegetation altitude change. With the atmospheric lapse rate of $5^{\circ}\text{--}6^{\circ}\text{ km}^{-1}$, the snow line and vegetation descent of about 1 km would imply temperature changes of this magnitude. The pollen data (Figs. 1 and 2) also imply that this cooling occurred at altitudes from 4.5 km down to at least 2 km.

CLIMAP SEA-SURFACE TEMPERATURES

CLIMAP sea-surface temperatures for the LGM are quantitative estimates based upon factor analysis of modern sediment data and the use of transfer functions to relate fossil assemblages to LGM temperatures. The quality (i.e., consistency and accuracy) of these estimates for the LGM have been extensively discussed, and the estimate of uncertainty at low latitudes is 1°C (CLIMAP Project Members, 1976, 1981; Prell *et al.*, 1980; Molino *et al.*, 1982; Prell, 1982).

CLIMAP sea-surface temperature reconstructions are given for February and August. As we desired to integrate a general circulation model for the full annual cycle,

it was first necessary to produce values of the sea-surface temperatures for each month. To do this the CLIMAP sea-surface temperatures for February and August were taken as the (mid) monthly extremes in each hemisphere. Comparison with current sea-surface temperatures indicated that a simple sine curve, with these extremes, matched to a first order the observed annual cycle. A sine curve was thus fit to the CLIMAP February and August values to produce 18,000 yr B.P. sea-surface temperatures for each month. The amplitude of the sine curve is small at low latitudes and introduces marginal uncertainty. The fit was made using the $2^{\circ} \times 2^{\circ}$ resolution of the data, then area weighted to the model's $8^{\circ} \times 10^{\circ}$ grid. In practice, the sea-surface temperatures at each grid point are updated each day to provide a smooth annual and monthly progression. Figure 3 shows the sea-surface temperature anomaly (18,000 yr B.P. minus current) used in the model for February, August, and the annual average.

Sea-surface temperatures at low and subtropical latitudes in the CLIMAP reconstruction are somewhat similar to modern values (Fig. 3). The subtropical gyre regions in the Pacific are actually warmer than in the modern climate; near Hawaii, surface temperatures are about 2°C warmer. In the other regions of interest, annual average sea-surface temperatures are 2°C colder off New Guinea and Colombia and 1°C colder off East Africa.

The 1981 CLIMAP sea-surface temperature reconstruction in the Pacific involved reassessment of data and resulted in warmer temperatures than in the 1976 reconstruction. In particular, the 1976 reconstruction resulted in sea-surface temperatures 3°C colder than today during Northern Hemisphere summer between 0° and 10°S (Gates, 1976); the 1981 reconstruction gives a cooling of only 1.5°C . As shown below, this modification makes it harder to reconcile CLIMAP and terrestrial data at low latitudes.

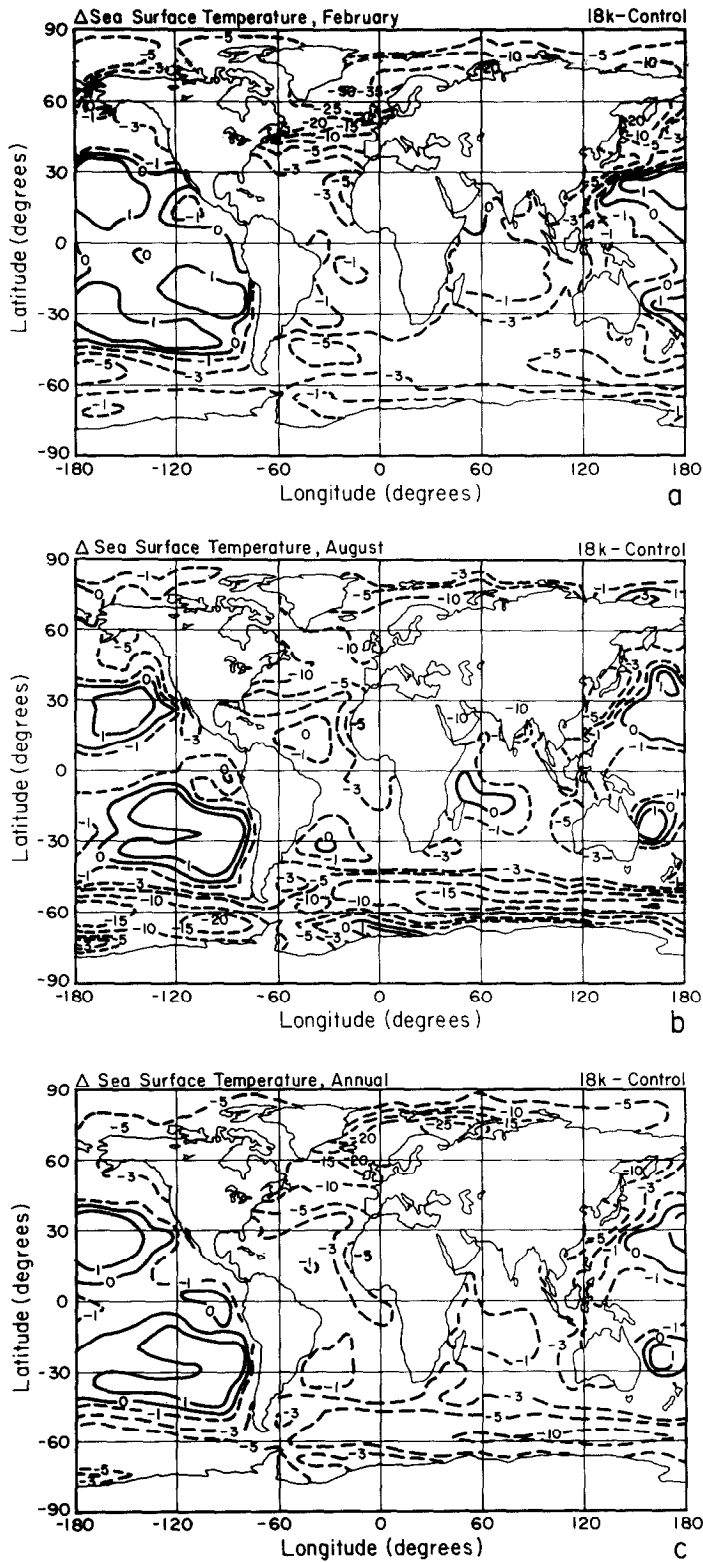


FIG. 3. Change in sea-surface temperature (or the surface temperature of sea ice) between 18,000 yr B.P. and the present: (a) February, (b) August, and (c) annual mean.

GENERAL CIRCULATION MODEL

The three-dimensional global climate model used for these experiments is described in Hansen *et al.* (1983, model II). The model has realistic topography and an $8^\circ \times 10^\circ$ (lat \times long) horizontal resolution, with fractional grid representation for land and ocean. There are 9 vertical levels, with a top at 10 mbar. The model solves the equations for conservation of mass, energy, momentum, and moisture; radiation includes all significant atmospheric gases, aerosols, and cloud particles, with cloud cover being predicted. Ground temperature calculations include diurnal variation and seasonal heat storage, while ground hydrological parameters are a function of vegetation type. As described in Hansen *et al.* (1983), the model produces generally realistic temperature and precipitation fields, while perhaps its chief deficiency is the underestimation of the Ferrel Cell and associated transports.

A comparison of model-generated values with observed temperatures, precipitation, and evaporation is given in Table 2. Observations come from Crutcher and Meserve (1970), Korzoun (1977), Rudloff (1981), and Thompson (1965), as well as those cited by Porter (1979) and Webster and Streten (1978). The model temperatures are accurate in all locations, aided by the specification of the proper sea-surface temperatures from climatology. The rainfall is also well simulated except in East Africa, where the model value is appropriate only for the northern sections of the grid box. In this area, in both the model and in reality, there are large gradients in rainfall totals, and the model simulations of 200 mm in the grid to the east and 1500 mm in the grid to the west are accurate. Evaporation values appear too large over land; however, evaporation is very difficult to measure accurately. The monthly variation of the surface and upper wind directions are shown in Figures 7–10. The only deviation from observations occurs in New Guinea, where the summer Walker circulation extends too far east in

the model. Otherwise, the general circulation and synoptic features are realistic [see Hansen *et al.* (1983) for a complete comparison of model simulations with numerous atmospheric observations].

To apply this model to the LGM, the CLIMAP boundary conditions (CLIMAP Project Members, 1981; Denton and Hughes, 1981) of sea-surface temperature, land ice, and sea ice were employed, along with the appropriate orbital parameters (Berger, 1978). The sea-ice values for February and August were taken to represent extremes; linear interpolation between them provided values for the intervening months. Land ice was assumed constant throughout the year. While the CLIMAP boundary conditions also provide estimates of ground albedo, the modern values based on vegetation type were used in this experiment for the unglaciated regions. An alternate run with vegetation changed to be consistent with the Köppen classification scheme produced no change in the values of interest here (due to the dominant influence of sea-surface temperatures). We note that when Manabe and Hahn (1977, Fig. 24) changed all of the land albedos back to modern values (involving removal of glaciers and changing glacial albedos to non-glaciated values) they found that the effect on low-latitude precipitation was relatively small ($\sim 20\%$ of the total change) everywhere except over the Indian subcontinent. With the fall in sea level due to the accumulation of ice on land, the area of exposed land increased by 4%; with the buildup of continental ice sheets, the mean altitude of the land increased by 115 m. The average annual sea-ice cover increased an amount equal to 4% of the area of the globe.

To test the sensitivity of the tropical continental temperatures and precipitation to the sea-surface temperatures, an alternate LGM model experiment was run which differed from the standard run in having sea-surface temperatures reduced everywhere by 2°C . In all other respects the two LGM runs had identical initial and boundary conditions.

The control run and both LGM runs were integrated for 6 yr starting from model-generated initial conditions. The first year allows the model to adjust to the boundary and initial conditions; the results presented below are averages for years 2–6.

GENERAL CIRCULATION MODEL RESULTS

The global average surface temperature change (averaged over land and ocean) determined by the model is -3.6°C (Fig. 4a). To analyze the contributions to the global cooling from the various parts of the climate system, the LGM model was rerun with different boundary conditions replaced with current values one at a time (e.g., land ice, sea ice, vegetation), allowing feedbacks like cloud cover and water vapor to vary, and also keeping them fixed (Hansen *et al.*, 1984). This analysis produced the following results: the contribution to the global cooling from the change in land ice, including both the effect of the elevation change and the change in albedo, is approximately -0.9°C , from the increased sea ice approximately -0.7°C , from decreased water vapor (due to the colder climate producing less evaporation) approximately -1.3°C , and from the cloud cover change (increased low-level clouds at midlatitudes, and decreased cirrus clouds) approximately -0.7°C .

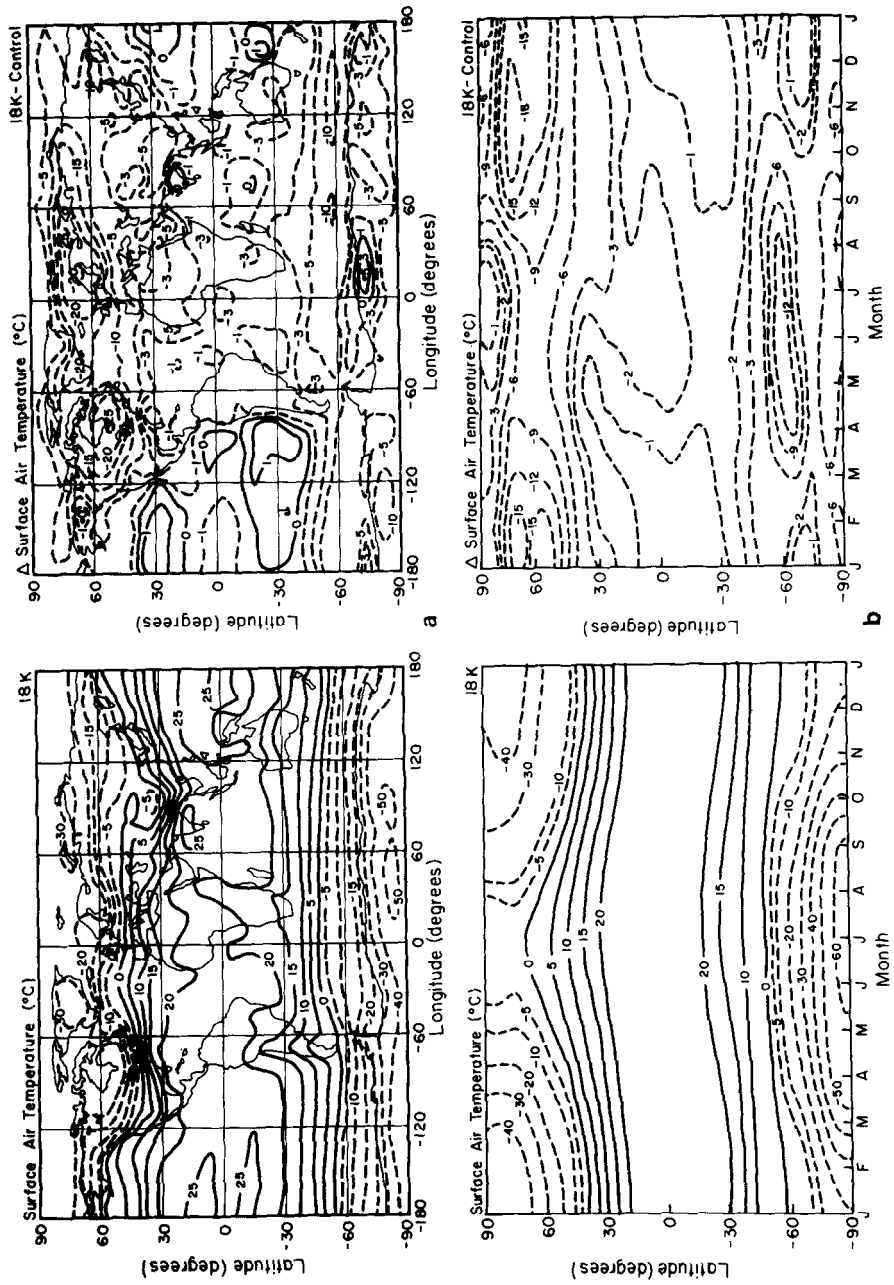
The high-latitude change is most evident in winter (Fig. 4b), as the greater atmospheric stability isolates the change near the surface. In the equatorial region the temperature change is small throughout the year, as was the change in CLIMAP sea-surface temperatures. At low latitudes there is a fairly consistent cooling of about 2°C throughout the lower troposphere (Fig. 4c), implying that the lapse rate does not change greatly.

The equatorial alpine paleoclimatic data indicate a snow line depression to altitudes around 3500 m during the LGM. In the model simulation, however, at low latitudes the 0°C isotherm is still found at 4.5 km

(Figs. 5a, b), temperatures having cooled about 2°C in most low-latitude regions.

The relatively warm subtropical and tropical sea-surface temperatures from the CLIMAP reconstruction come in contact with cold continental air during Northern Hemisphere winter, leading to increases in evaporation of 20–50%. The trade winds advect this moisture toward the equator where it leads to an increase in rainfall (Fig. 6). Hansen *et al.* (1984) provide additional discussion of the feedback mechanisms for the model's LGM climate while Rind (1985) discusses the atmospheric dynamics.

We concentrate here on the results for the four individual locations. Annual averages are given in Table 2. The LGM model results for Hawaii are from the grid box which extends from 16° to 24°N , 155° to 165°W . Note that Hawaii represents only a small portion of this grid box, and orographically induced precipitation or mountain-top temperatures cannot be simulated. LGM surface air temperature is actually slightly warmer in winter and spring (Fig. 7), a result of the CLIMAP annual average sea-surface temperatures being 2°C warmer in this grid box (CLIMAP, 1981). These warm temperatures lead to increased convection and precipitation, except during the summer. In the model, convection mixes momentum and the increased convection weakens the trade wind inversion, thereby decreasing wind speeds at upper and lower levels. As evaporation shows little change, the ground becomes wetter. The surface wind direction becomes more southeasterly, a result of a changed sea-level pressure distribution. CLIMAP sea-surface temperatures indicate a greater cooling farther east in the eastern Pacific at Hawaii's latitude and this, in association with the increased precipitation near the equator in the eastern Pacific (Fig. 6), helps generate a strengthened local meridional cell-rising motion near the equator, and a sinking motion at about 20°N . The result in the sea-level pressure field is a strengthened subtropical high to the south and east of its previous position. The position of this sub-



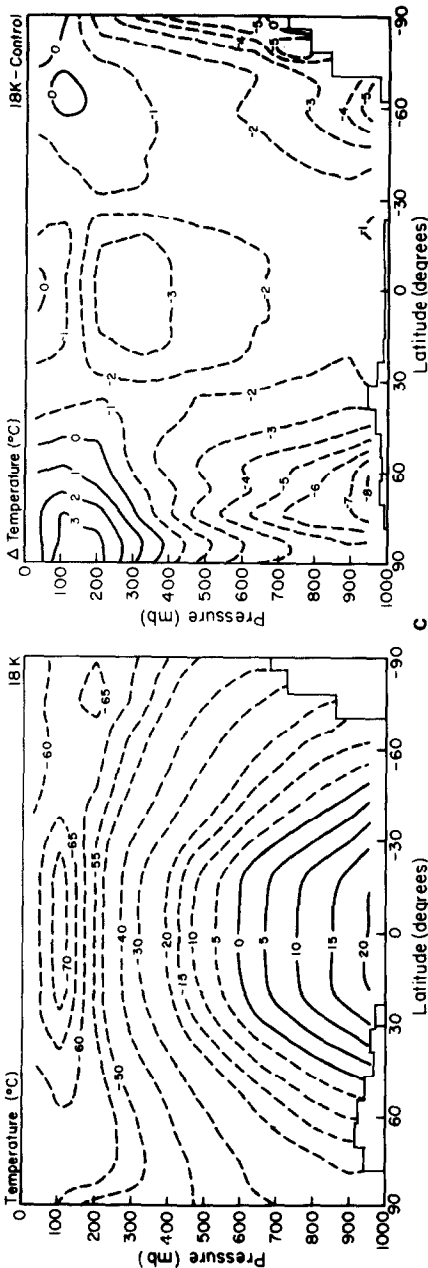


FIG. 4. Surface air temperature for the 18,000 yr B.P. experiment (left) and temperature change from the current climate control run (right) from years 2–6 for (a) latitude–longitude for the annual average, (b) latitudinal average versus month (middle), (c) latitude–altitude profile of temperature and temperature change for the annual average.

tropical high results in the more southeasterly surface wind direction shown in Figure 7. The annual average surface air temperature is higher by 0.5°C , significant at the 95% level using the two-sided t test. The annual precipitation is 550 mm greater (95% significance), and the increase in October and November cloud cover is significant (95%).

The LGM model results for Colombia are for the grid box that extends from 0° to 8°N , 65° to 75°W . Reconstructed CLIMAP sea-surface temperatures in this region are 2°C colder for the annual average, with changes of close to -5°C in summer. Surface air temperature variations are similar (Fig. 8), although up to 1°C of this change may be due to the slightly greater topography in the CLIMAP data set for this grid box. Maximum cooling and decreased rainfall thus occur during summer. The low-latitude precipitation increase due to increased moisture advection is seen from March through May, when offshore cooling is least. Thus, while the movement of the ITCZ still maintains maximum precipitation during the equinoxes, the sea-surface temperatures introduce an asymmetry, with the spring equinox being rainier. As the cold sea-surface temperatures during summer help stabilize the air over northwestern South America, maximum precipitation shifts to the northeast. This builds the sub-equatorial ridge at higher levels into this region, and upper-level winds shift to the east or southeast. The annual average surface temperature is 2.5°C colder (99% significance) (Table 2) and the April and May precipitation increases are significant at the 95% level.

The model results for East Africa are for the grid box from 0° to 8°N , 35° to 45°E . The LGM annual sea-surface temperature is about 1°C cooler offshore to the east. Surface air temperature is generally 3° – 4°C cooler (Fig. 9), with perhaps 1° – 2°C of this due to the different topography for this grid in the CLIMAP data set, and precipitation increases with little change in evaporation.

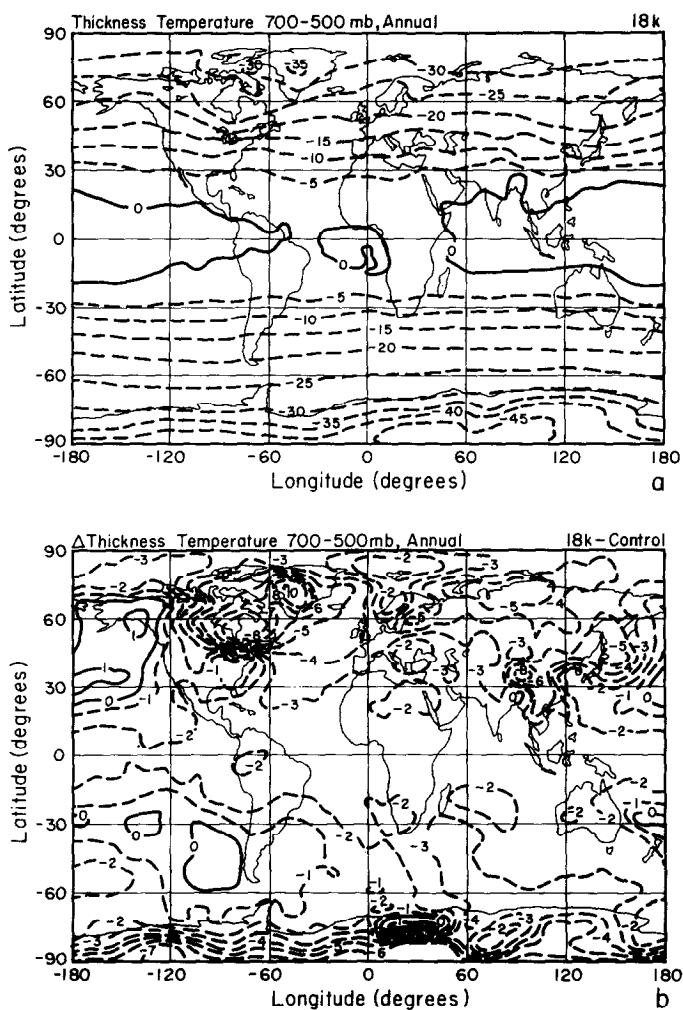


FIG. 5. (a) Annual mean (thickness) temperature of the 700–500 mbar layer, which represents 4.5 km altitude from 30°N to 30°S in the 18,000 yr B.P. experiment and (b) change in this temperature, 18,000 yr B.P. climate – control.

The seasonal character of the precipitation change closely resembles the latitudinal average (Fig. 6). The colder Arabian continent strengthens the Arabian high in winter, providing slightly more northerly and stronger surface winds. In addition, the high reestablishes itself more quickly in the fall (October), as seen in the return to northeasterly flow; the upper air flow also indicates a one-month-quicker return. The stronger high and increased low-latitude precipitation are indicative of an increased meridional cell at this longitude. For the annual average the surface temperature is

3.1°C lower (99% significance) and the precipitation is 220 mm greater (90%, and thus of marginal significance) than present values (Table 2).

The model results for New Guinea are from the grid box 0° to 8°S, 135° to 145°E. The LGM sea-surface temperatures in the vicinity are about 2°C cooler, with maximum cooling during May–September (Southern Hemisphere winter). Land to the south rises above sea level, and the land portion of this grid box increases from 55 to 70%. Figure 10 shows the change in climate parameters. The cooling and drying

are at a maximum in winter. At upper levels, intensified ridging over Australia provides for a greater south wind throughout the year; the altered longwave pattern in this region closely resembles the hypothetical reconstruction shown by Webster and Streten (1978). Annual precipitation for the western Pacific decreased by 580 mm, while in the eastern Pacific it increased by 270 mm. This is consistent with the overall cooling of the west Pacific sea-surface temperatures relative to those in the east (by 1.2°C). This result and the decreased rainfall in New Guinea are consistent with a weakened Walker circulation. The annual average surface temperature was 1.6°C lower (99% significance), and the precipitation was about 660 mm less (95% significance) than in the simulation of the present climate.

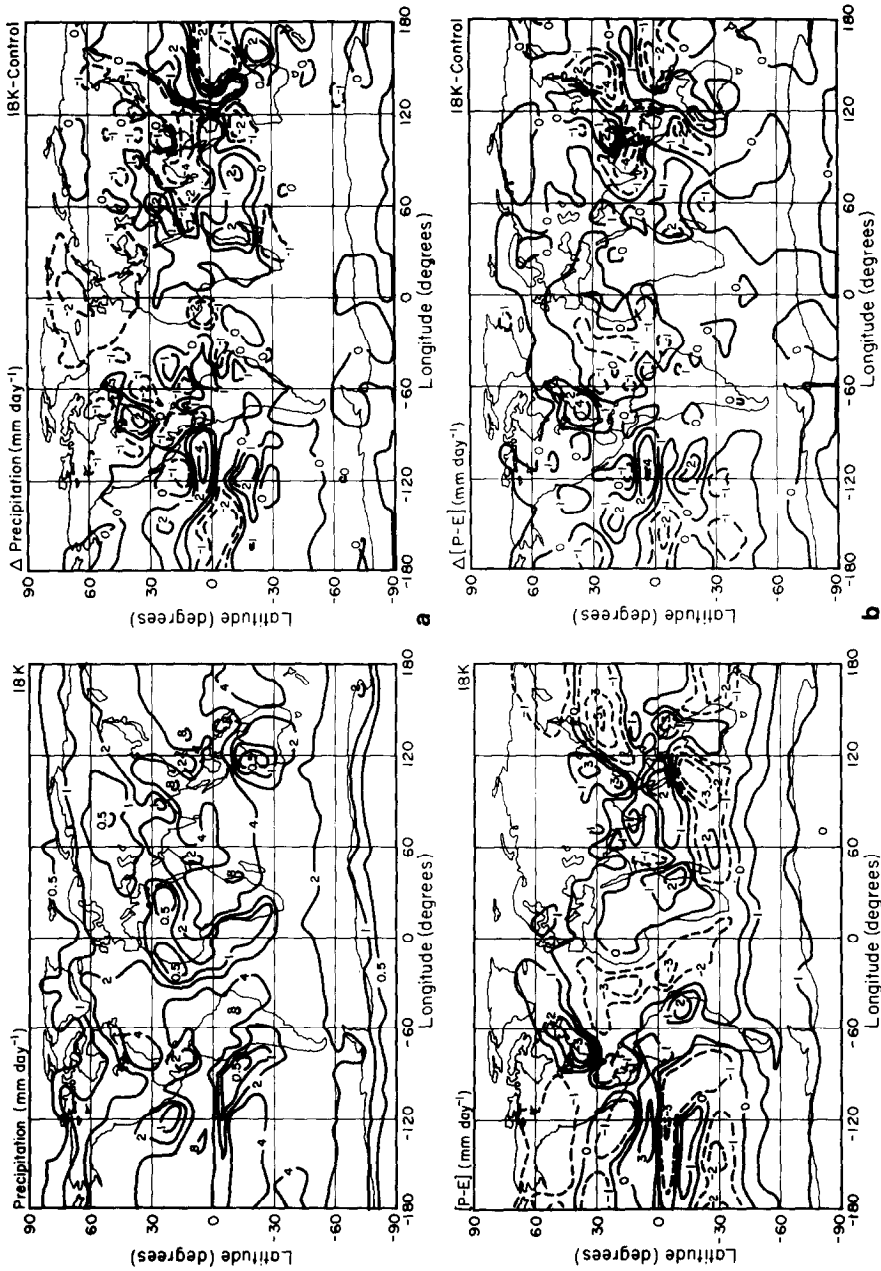
The annual average temperature profile in the present-day climate simulation crosses the 0°C isotherm between 4.5 and 5 km in all cases, in good agreement with the present-day snow line (Fig. 11) (Schubert and Medina, 1982). In the LGM run the 0°C isotherm descends about 500 m in East Africa, New Guinea, and Colombia, while showing no variation over Hawaii. Table 2 indicates that the first three locations experienced surface air temperature cooling of 1.5°–3.1°C, with sea-surface temperatures about 1°–2°C colder, while in Hawaii the surface air temperature was 0.5°C warmer and the sea-surface temperatures were 2°C warmer. In the areas outside Hawaii the descent of the freezing level is only about half that implied by the snow line data, while the cooling is, at most, half that indicated by the pollen data (Table 2, 2.5-km temperature), using our nominal palynological estimate of 5°–6°C cooling. While there is no cooling estimate available for Hawaii from the pollen data, the inferred snow line depression of 935 ± 190 m on Hawaii is unexplained by the temperature change as the model produced no descent of the freezing level.

COMPARISON WITH OTHER MODELS

These model results can be compared with previous GCM studies (Williams *et al.*, 1974; Gates, 1976; Manabe and Hahn, 1977). Besides the obvious differences between models (these use finer horizontal resolution, the GFDL model does not calculate clouds, etc.), they used sea-surface temperature data sets from older reconstructions. Gates, as well as Manabe and Hahn, used the 1976 CLIMAP sea-surface temperatures which, as noted, are colder than those reported in 1981; Williams *et al.* used an even earlier data set with colder values. The differences in both cases were mainly in the tropics and subtropics. As these runs were generally for Northern Hemisphere summer (e.g., July or August), we will compare results during this season. Williams and Barry (1975) found a global cooling much larger than their 5°C tropical cooling. Gates (1976) derived a global cooling of 4.9°C, and Manabe and Hahn (1977) computed a global cooling of 5.4°C. The contrast to the results obtained here (–3.5°C during June–August) is due mainly to the colder sea-surface temperatures used in those studies. This also affects other aspects of the simulations: the global average precipitation reduction obtained by Gates was 14%, while Manabe and Hahn obtained 10%; in particular, there was extensive drying at low latitudes over land for both models, consistent with paleoclimate evidence. In contrast, in our run the June–August precipitation was only 5% less, and, in general, there was little drying over land at low latitudes.

ALTERNATIVE LGM EXPERIMENT

To estimate the sensitivity of the model results to sea-surface temperatures, an alternative run was made, with ice-age boundary conditions identical except that all sea-surface temperatures were lowered by 2°C relative to the CLIMAP values. As shown in Figure 11, the 0°C isotherm now descends by about 1000 m in Colombia,



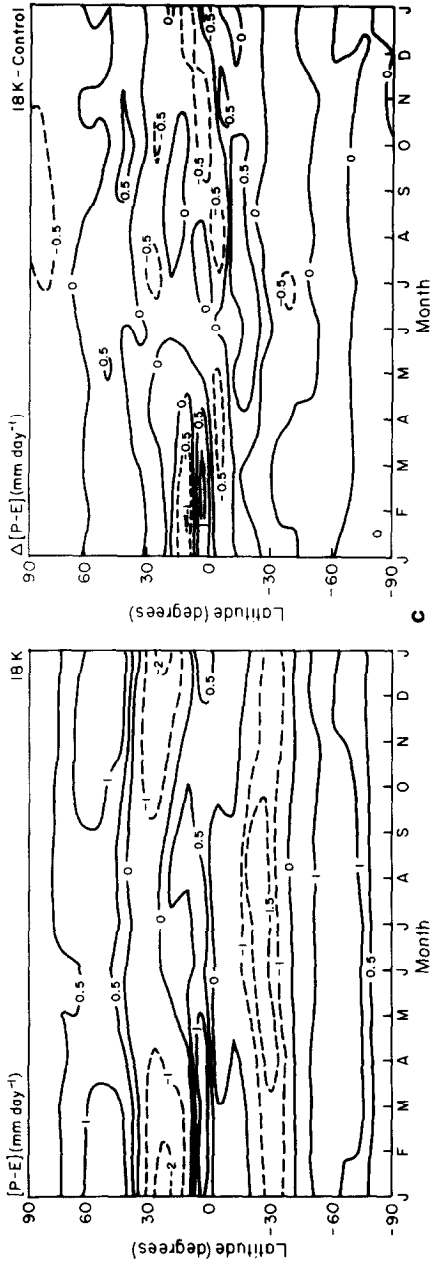


FIG. 6. Hydrologic cycle (left) and its change 18,000 yr B.P.—control (right) from years 2–6 for (a) annual average precipitation (latitude \times longitude), (b) annual average precipitation minus evaporation (latitude \times longitude), and (c) latitudinally averaged precipitation minus evaporation as a function of the month.

East Africa, and New Guinea, and by about 500 m in Hawaii. Global cooling is now 5.5°C during summer, precipitation decreases by 13%, and there is less precipitation over land at low latitudes. All of these results are in good agreement with the other models. The changes in the climate parameters at the four locations are shown in Table 2.

DISCUSSION

The results presented above indicate that the ice-age climate simulation using CLIMAP sea-surface temperatures does not provide sufficient cooling at the specific low- and subtropical-latitude locations to match the terrestrial pollen and snow line data, assuming that they indicate temperature depression at 18,000 yr B.P. (compare Tables 1 and 2). A companion experiment in which the CLIMAP sea-surface temperatures were reduced everywhere by 2°C is in better agreement with the land data, although for Hawaii the model freezing-level descent is still less than the observed snow line descent. These results raise several fundamental questions, and illuminate areas where further research is needed. Several of these issues are addressed below:

(1) Does this inconsistency concern modeling capabilities? Perhaps the marine and continental data sets are consistent and lapse rates simply increased greatly in mountain areas. Neither this model nor the GFDL model (with finer resolution) can resolve small-scale effects in mountain regions; both show only a small change in lapse rate (Manabe and Hahn, 1977, Fig. 17), consistent with the small change in moist adiabatic lapse rate. We return to this point below. To isolate mountain effects, additional modeling studies should use as fine a resolution as is practical.

The model also shows little change in rainfall or rainfall minus evaporation at low latitudes over land (Fig. 6 and Table 2). However, substantial evidence exists for worldwide aridity at low latitudes during

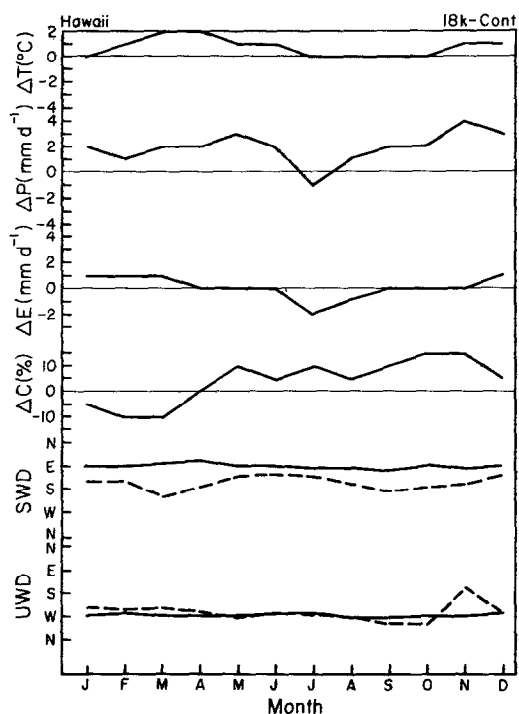


FIG. 7. 18,000 yr B.P. — control changes as a function of the month for the grid box which includes Hawaii. Changes shown are for surface air temperature, precipitation, evaporation, and cloud cover. Also shown are the surface air and upper air (200-mbar) wind directions, for both the control run (solid line) and the 18,000 yr B.P. experiment (dashed line).

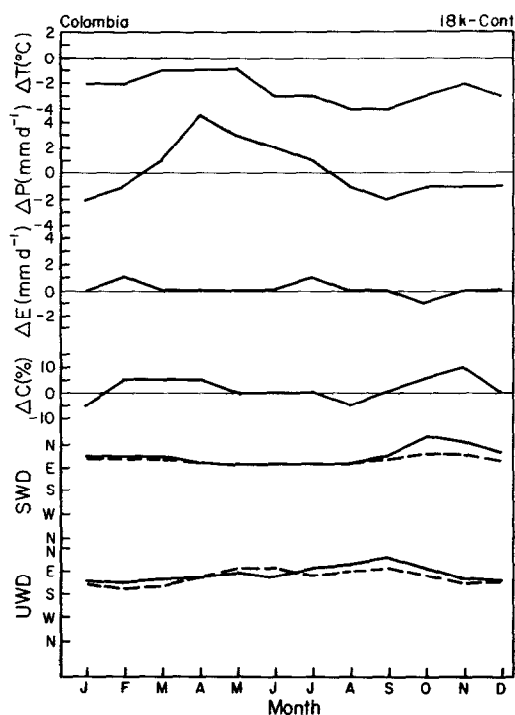


FIG. 8. As in Figure 7 for the grid box which includes Colombia, South America.

the LGM [e.g., Sarnthein's (1978) worldwide data summary of fossil dunes, indicating that LGM sand-dune formations and associated deserts were much more widespread than they are today; Street and Grove's (1979) results showing low lake levels in tropical Africa and South America (see also Peeters, 1984); pollen results from the Galapagos (Colinvaux, 1972); and the pollen data discussed above for several different locations which show increased aridity].

Probable contraction of the Amazon and African rain forest during the LGM is suggested by floral and faunal evidence (Vuilleumier, 1971; Hamilton, 1976; Flenley, 1979; Smith, 1982). When the Köppen climate classification scheme was applied to the temperature and precipitation results of

the standard 18,000 yr B.P. run, no such contraction appeared (Hansen *et al.*, 1984). The alternate model experiment resulted in substantial dessication of the rainforest both in South America and Africa, due to the combination of increased cooling and increased aridity (the alternate 18,000 yr B.P. run resulted in more than double the decrease in the hydrologic cycle compared with the standard 18,000 yr B.P. experiment). Thus the intensity of the hydrologic cycle at low latitudes in the atmospheric models is strongly a function of the low-latitude sea-surface temperatures, as exemplified by the greater aridity produced by all of the models when colder LGM sea-surface temperatures were used. This dependency is reasonable, for evaporation, and thus precipitation, should decrease as ocean temperatures cool, but the proper magnitude of the relationship is hard to establish. A focused research question is: could low-latitude continents experience substantial drying if there was little change

TABLE 2. ANNUAL AVERAGE OBSERVATIONS AND MODEL-GENERATED RESULTS FOR THE CURRENT CLIMATE, 18,000 yr B.P. SIMULATION, AND ALTERNATE 18,000 yr B.P. SIMULATION

	Precipitation (mm)	Evaporation (mm)	Nearby sea-surface temperature (°C)	Surface air temperature (°C)	2.5-km temperature (°C)	Freezing level (m)	Surface 4-km lapse rate (°/km)	Moist adiabatic lapse rate (°/km at 2 km)
Hawaii								
Observed	1200	2000	25	24	11	4700	5.5	4.6
Current	1450	1820	25	24	11	4500	5.5	4.6
18,000 yr B.P.	1998	2002	27	24.5	11	4500	5.6	4.6
Alt. 18,000 yr B.P.	1779	1892	25	22.6	9.5	4000	5.7	4.8
Required ^a							7.3 (6.9)	
Colombia								
Observed	2400	1250	25	24	12	5000	5.3	4.5
Current	2482	2117	25	24	12	4900	5.3	4.5
18,000 yr B.P.	2592	2154	23	21.5	11	4500	5.4	4.6
Alt. 18,000 yr B.P.	2373	2044	21	19.4	9	4000	5.7	4.8
Required ^a							6.4 (6.7)	
East Africa								
Observed	1200	700	26	21	14	4850	5.0	4.3
Current	2263	1825	26	21	14	4950	5.0	4.3
18,000 yr B.P.	2409	1934	25	17.9	11.5	4500	5.3	4.6
Alt. 18,000 yr B.P.	2299	1824	23	15.7	9	3950	5.6	4.9
Required ^a							7.1 (6.3)	
New Guinea								
Observed	4000	1400	29	24	14	4700	5.6	4.3
Current	4307	2190	29	24	13.5	4900	5.6	4.4
18,000 yr B.P.	3650	2299	27	22.4	11	4400	5.9	4.7
Alt. 18,000 yr B.P.	3212	2189	25	20.3	9	3900	6.0	4.9
Required ^a							7.5 (6.8)	

^a Calculated from the difference between the sea-surface temperature and the ice-limit data; the value in parenthesis is from the 18,000 yr B.P. model surface air temperature and ice-limit data.

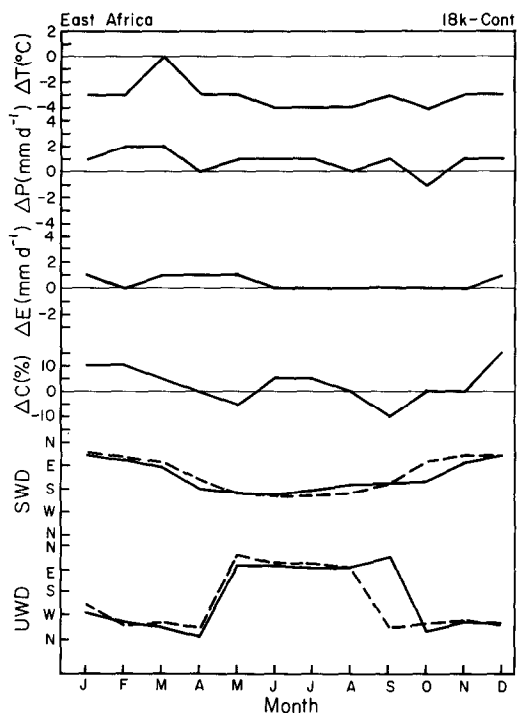


FIG. 9. As in Figure 7 for the grid box which includes East Africa.

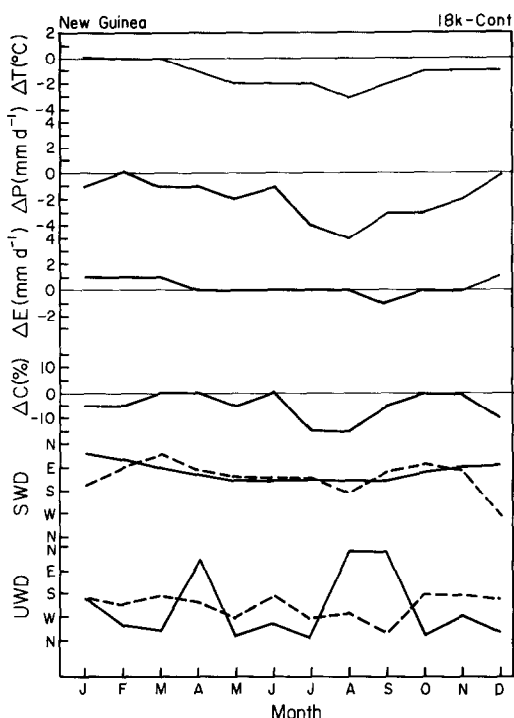


FIG. 10. As in Figure 7 for the grid box which includes New Guinea.

in low- and subtropical-latitude ocean temperatures?

(2) Could an LGM lapse rate change necessitated by the coexistence of warm sea-surface temperatures and montane temperature depression occur throughout the tropics with very little change in the moist adiabatic value? Consistency between the snow line data and CLIMAP sea-surface temperatures would require large increases in lapse rates (Table 2, "required" value), with even larger values needed to accommodate the inferred temperature changes at 2–3 km based on pollen data. The lapse rate change would have to be widespread; in addition to the four widely dispersed regions discussed here, mountain glaciers also expanded in Mexico, other areas of South America, and possibly in Central America and the Greater Antilles (Mercer and Palacios, 1977; Hastenrath, 1981, Table 5; Denton and Hughes, 1981; Schubert and Medina, 1982, Table 1). At low latitudes the current lapse rate is consistently close to

the moist adiabatic value (Table 2; Webster and Streten, 1978; Stone and Carlson, 1979) and the moist adiabatic value would not have shown much change if sea-surface temperatures remained warm during the LGM. Current understanding of atmospheric dynamics suggests that a large divergence of the tropical LGM lapse rate from the moist adiabatic value is implausible, as moist convection should still represent the dominant vertical heat-transporting process at low latitudes. Webster and Streten (1978) concur and show that at low latitudes even arid stations have lapse rates close to moist adiabatic.

(3) How reliable are the snow line data, and their interpretation? As shown in Table 1, the glacial chronologies are only broadly bracketed by dates which, while encompassing the LGM, do not confirm the extent of the glaciers at 18,000 yr B.P. Pollen data (Figs. 1 and 2) do not support the idea that temperatures were colder at some time other than 18,000 yr B.P. nor do they indi-

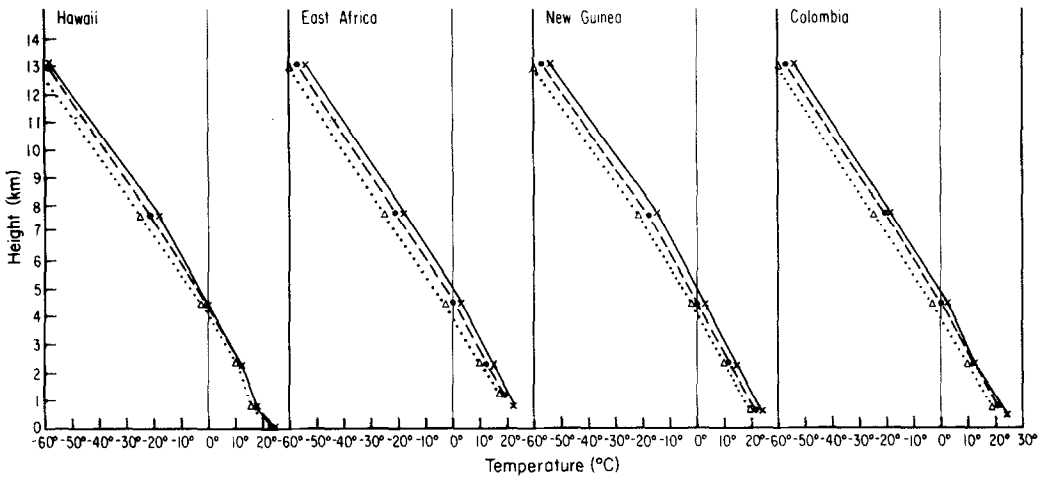


FIG. 11. Annual mean vertical temperature profile for Hawaii, East Africa, New Guinea, and Colombia for the control run (solid line), 18,000 yr B.P. experiment (dashed line), and alternate 18,000 yr B.P. experiment (dotted line).

cate large fluctuations near 18,000 yr B.P. We must, however, allow for the possibility that glaciers reached their maximum at some time(s) other than 18,000 yr B.P., which might imply that sea-surface temperatures were colder at this alternate time.

It has been suggested that snow line depression could result in part from an increase in precipitation. Model results do show increased precipitation over Hawaii with southeast winds, along with a weakening in the trade wind inversion (as suggested by Porter, 1979), and the results for New Guinea support increased excursion of air from higher latitudes, possibly bringing more frequent snow, as suggested by Webster and Streten (1978). Under average cloud conditions, and with a 70% albedo for freshly fallen snow, about 100 W m^{-2} is absorbed at the surface at low latitudes. This would melt 10 cm of snow (1-cm water equivalent) in about 9 hr; if the air temperature is 2°C above freezing, additional fluxes of sensible and latent heat would reduce the melting time to 6 hr. Continual replenishment of snow would be necessary to maintain snow cover, which would require frequent snowstorms, with subfreezing temperatures; thus, glacier equilibrium lines, dividing the regions of

net snow accumulation and ablation, would tend to be near the 0°C isotherm on average (as it is now in the tropics: Butzer, 1971, p. 109). Due to the high albedo of fresh snow, an increase in cloud cover would not drastically alter this assessment. Further research and more detailed calculations are needed to establish the conditions necessary to maintain tropical glaciers substantially below the freezing level for extended periods of time [comparable to the study of Porter (1977b) for the Cascade Range in Washington].

(4) What caveats are involved in estimates of temperature depression derived from pollen changes? It is important to emphasize that such estimates do not result from the type of factor analysis used to provide paleosea-surface temperature estimates; they result from the combination of measured altitude changes with prescribed lapse rates. In addition, there is a need to distinguish between the effects of cooling and drying, and the possible effects of reduced CO_2 on tropical vegetation. Nevertheless, the pollen-derived cooling estimates cannot be dismissed; the results shown in Figures 1 and 2 indicate widespread depression of vegetation zones during the LGM. More systematic studies

of the relationships between modern pollen and vegetation and between vegetation and temperature are needed, as well as additional pollen data with radiocarbon control throughout tropical areas before one can assign uncertainties to the estimated temperature decreases (e.g., Table 1).

(5) How accurate are the CLIMAP sea-surface temperature reconstructions? The CLIMAP data set represents a compilation, with comprehensive factor analysis, which has been published and scrutinized. The uncertainties given for the CLIMAP temperature reconstructions are not sufficient to allow the model to reproduce the terrestrial data, either for air temperature (assuming the terrestrial data represents temperature depressions of 5° – 6° C at 18,000 yr B.P.) or precipitation. If errors exist in the data set, they might be in regions with few cores. One such location is the subtropical North Pacific where the reconstruction produces very warm subtropical gyres; could these really have been maintained in a much colder climate? Obviously, more cores with fossil preservation are needed. More critically, there is no way to prove the basic assumption that the long-term change (or lack of change) in fossil assemblages is directly translatable to sea-surface temperature changes. This data set is much better quantified than the terrestrial data set [level II compared with level I for the terrestrial data in the terminology of Peterson *et al.* (1979)], but the terrestrial data seem to be producing a similar picture at widespread locations (Table 1).

(6) Finally, were the ocean temperatures coldest at 18,000 yr B.P.? Recent sea-surface temperature reconstructions based on tropical Atlantic foraminifera show both spatial and temporal variability, with minimum ocean temperatures occurring in different regions at different times (Mix *et al.*, 1983). Thus the CLIMAP reconstruction for 18,000 yr B.P. may not indicate the coldest full-glacial sea-surface temperatures for all locations. If so, temperatures over land in certain areas could have been

colder at some time other than 18,000 yr B.P. Although more precise dating is needed, the vegetation depression in widely dispersed regions does not suggest greater cooling at times other than 18,000 yr B.P. If the CLIMAP sea-surface temperatures in many regions do not represent the coldest ocean temperatures of the LGM, then it may be misleading to use this data set to reconstruct the coldest climate of the last ice age.

The LGM model experiment, with specified sea-surface temperatures, was not in radiation balance; if the sea-surface temperatures had been allowed to change, they would have cooled by an additional 1.2° C (the cooling effect of water vapor decrease and cloud cover change was mitigated by fixing the sea-surface temperatures; Hansen *et al.*, 1984). This cannot be considered proof of a warm bias in the CLIMAP temperatures because of the uncertainty involved in modeling cloud cover; however, if the CO_2 had been reduced to 240 ppm, as suggested by ice-core data (Shackleton *et al.*, 1983), and the vegetation changed, the resulting climate (without specified sea-surface temperatures) would have been 5° – 6° C colder than at present. There is a substantial difference between the climate sensitivity necessary to provide a 5° – 6° C cooling and the 3.6° C change which results from using the CLIMAP sea-surface temperatures. As the questions of climate sensitivity and of the relative sensitivity of low and high latitudes are central to the estimation of the rapidity and size of future climate changes, the topic deserves especially close scrutiny.

ACKNOWLEDGMENTS

Special thanks go to Dr. W. Broecker for suggesting this investigation, and Dr. A. Lacis and Dr. J. Hansen for their analysis of the 18,000 yr B.P. radiation balance. We thank A. Mix, R. Seager, Dr. C. J. Heusser, and Dr. L. Heusser for their comments on the manuscript. We also express appreciation to H. Brooks, L. Smith, K. Prentice, Dr. R. Ruedy, and Dr. G. Russell for assistance with the 18,000 yr B.P. experiment, J.

Lerner for computer plots, J. Mendoza and L. Del Valle for drafting, and C. Paurowski and E. Michaud for typing.

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